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Tropospheric Stratospheric
Turbulence and Vertical Diffusivities

S. P. ZIMMERMAN
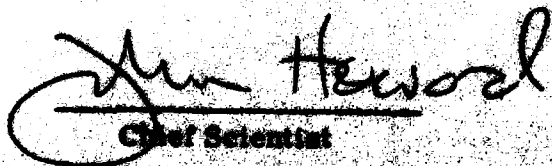E. A. MURPHY

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The determination of winds and temperature by rawinsonde allows us to calculate the Richardson number, a measure of atmospheric instability. From this, we can then determine the turbulent atmospheric parameters, diffusivity (K), rate of heat deposition (ϵ), and the vertical turbulent intensity ($<w^2>$). Comparison of these data, for the spring periods, with contours of residual stratospheric radioactivity demonstrate that, at the tropopause, the isocountours of turbulent diffusivity accurately follow the isodensity contours of the residual radioactivity. We also demonstrate that the rapid

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cessation of the turbulent diffusivity accurately predicts the level of the tropopause.

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Tropospheric Stratospheric Turbulence and Vertical Diffusivities

I. INTRODUCTION

The utilization of radioactive tracers deposited into the stratosphere to deter-
mine effective stratospheric-transfer coefficients is well documented by Reed and
German, [1] Gudiksen et al, [2] Luther, [3] and Danielsen and Louis. [4] In these studies,
the dispersion of tracers in the stratosphere and their rate of emission from the
stratosphere was used to determine "effective" stratospheric diffusion coefficients.
It is considered that in these two-dimensional analyses the longitudinal diffusivity
(K_{yy}) is due to the effect of mean meridional motions and large wave transfer. The
tilted K_{yz} coefficients are due to the tilt of the tropopause when the transfer is
downward, from the equatorial tropopause to the polar regions. K_{zz}, the vertical

1. Reed, R. J., and German, K. E. (1965) A contribution to the problem of
   stratospheric diffusion by large-scale mixing, Mon. Weather Rev.
   93:313-321.

2. Gudiksen, P. H., Fairhale, A. W., and Reed, R. J. (1968) Roles of mean
   meridional circulation and eddy diffusion in the transport of trace substance
   in the lower stratosphere, J. Geophys. Res. 73:4461-4473.

3. Luther, F. M. (1973) Monthly Mean Values of the Eddy Diffusion Coefficient in
   the Lower Stratosphere, AIAA Paper No. 73-498.

4. Danielsen, F. F., and Louis, J. F. (1977) Transport in the stratosphere, in
   The Upper Atmosphere and Magnetosphere, Studies in Geophysics, NAS,
   Washington, D. C., pp 141-155.

diffusivity term, which has the smallest magnitude, is the term that completes the vertical transfer to account for the observations.

The questions we address in this report are: what is the function of troposphere-stratospheric turbulence in the vertical transfer of trace elements, mass, heat, and momentum, and how does it relate to the above-deduced parameters? To examine these questions, we have utilized a large body of "rawinsonde data" supplied by the Environmental Data Service National Climatic Center in Asheville, North Carolina. This data consists of winds, temperature, and pressure as a function of altitude obtained from all available stations (currently 144 stations) for the period 1948–1976. At the present time only the 1970–1976 data have been reformatted for processing on the CDC 6600 computer system. These twice-daily data are used in calculating the Richardson number, a stability criteria for determining the presence or absence of atmospheric turbulence, and is given by the following relationship:

$$R_i = \frac{g}{T} \left[\frac{\partial T}{\partial Z} + \Gamma \right] \Big/ \left[\frac{\partial V}{\partial Z} \right]^2 \tag{1}$$

where

g is the acceleration of gravity

$\partial T / \partial Z$ is the vertical temperature gradient

Γ is the dry adiabatic lapse rate (= 9.8 K/km)

and

$$\left[\frac{\partial V}{\partial Z} \right]^2 \equiv \left[\frac{\partial V_x}{\partial Z} \right]^2 + \left[\frac{\partial V_y}{\partial Z} \right]^2 \quad \text{is the square of the vertical shear}$$

of the horizontal winds.

The rawinsonde system is not ideally suited for taking measurements that will provide thermodynamic data at the required sampling intervals for this application. The data, as used in this study, were originally intended to provide northern hemisphere pressure maps for a number of millibar pressure levels. In order to determine the Richardson number at equal-height intervals, the temperature and wind-component data points are fit, using a Hermite interpolation algorithm.[5] The temperature function along with the derivatives of the winds and temperature, with

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5. Tsipouras, P., and Cormier, R. V. (1973) Hermite Interpolation Algorithm for Constructing Reasonable Analytic Curves Through Discrete Data Points, AFCRL-TR-73-0400, AD 768669.

respect to altitude, are interpolated at 100-m intervals and used to calculate the Richardson number. The probability of occurrence of turbulence is obtained by defining a critical Richardson number ($R_i = 1/4$ according to current literature) and assuming that turbulence is present when the Richardson number is equal to or less than this value.

This approach follows that of Zimmerman and Murphy[6] in a previous article wherein they examined the microscale transfer coefficients in the mesosphere utilizing rocket grenade data. Once given the Richardson number, they related the ratio of the vertical turbulent velocity (w) divided by the mean wind (V) to this parameter, following the empirical relation fitted to Deacon's analyses:[7]

$$\frac{w}{V} = \left\{ \begin{array}{l} -0.15 \ (R_i > 0) \\ 0.15 \ (R_i < 0) \end{array} \right\} \sqrt{R_i} + 0.08 . \tag{2}$$

Thus, given the Richardson number and the horizontal wind velocity, we assume that the ratio of a turbulent parameter to its mean quantity measured in the boundary layer may be scaled for use in the higher atmosphere. This a priori assumption appears validated by the resulting amplitudes in the mesosphere, and it was more recently supported, in absolute value, by Mansen et al.[8] Thus, given the above turbulent intensity $<w^2>$ and atmospheric stability, we may then estimate the unknown turbulent parameters described below.

2. STRATOSPHERIC AND MESOSPHERIC TURBULENCE

The assumptions used in the following analysis is that once we are given the turbulent intensities $<w^2>$ from Eq. (2) we can assume that the spectrum of the turbulent motions will be inertial and, as discussed below, the vertical length scale will be limited by buoyance. Thus, given inertiality and one dimensionality, we have

$$<V_B{}^2> = \int_{k_B}^{\infty} E(k)\ dk \tag{3}$$

and

6. Zimmerman, S.P., and Murphy, E.A. (1977) Stratosphere and mesospheric turbulence in Dynamical and Chemical Coupling, D. Reidel, Dordrecht, Holland, pp 35-47.

7. Deacon, E.L. (1959) The problem of atmospheric diffusion, Int. J. Air Pollut., pp 92-108.

8. Manson, A.H. (1980) to be published.

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$$E(k) = \alpha \epsilon^{2/3} k^{-5/3}$$

where $<v_B^2> \equiv <w^2>$ is the buoyancy-limited, vertical turbulent intensity set identically equal to $<w^2>$, as given in Eq. (2); $E(k)$ is the turbulent energy spectrum; α is the Kolmagoroff constant, and for the vertical dimension it has the value ~ 0.8; k is the wavenumber; and ϵ is the rate of turbulent dissipation. Thus, from Eq. (3) we derive

$$<w^2> \approx \frac{3}{2} \epsilon^{2/3} k_B^{-2/3} . \tag{4}$$

In a review of the Lumley-Shur theory of turbulence in a buoyant stably stratified medium,[9] it is stated that the turbulent spectrum will be inertial in the high wavenumber region, and that the "buoyant" subrange will strictly be a function of atmospheric stability in the form of the Brunt Vaisalla frequency (N) and wavenumber (k). As a consequence of this, ϵ, which is conserved in the inertial region, is a function of wavenumber in the buoyant region. Thus, the spectral energy is given by

$$E(k) = C_1 N^2 k^{-3} + \alpha \epsilon^{2/3} k^{-5/3}$$
$$= \alpha \epsilon^{2/3} \left[1 + \left(\frac{k_B}{k} \right)^{4/3} \right] k^{-5/3} , \tag{5}$$

where k_B is the transition wavenumber

$$\left\{ k_B = \left(\frac{C_1 N^2}{\alpha \epsilon^{2/3}} \right)^{3/4} \right\}$$

between the inertial and buoyant subranges. In this report, we presume that the vertical turbulent spectrum is "inertial" and limited in spectral extent to the transition wavenumber (k_B), and, thus, the vertical turbulent kinetic energy as given by Eq. (4) becomes

$$<w^2> = \frac{3}{2} \left(\frac{\alpha^{3/2} \epsilon}{C_1^{1/2} N} \right) . \tag{6}$$

9. Phillips, O. M. (1967) On the Bolgiano and Lumley-Shur theories of the buoyancy subrange, in Atmospheric Turbulence and Radio Wave Propagation, Nauka, Moscow, USSR, pp 121-128.

For $\alpha = \sim 0.8$ and $C_1 \rightarrow 2$ to 2.5 (for simplicity say 2.5), we have

$$< w^2 > \cong \frac{1}{2} \frac{\epsilon}{N} \qquad (7)$$

or

$$\epsilon \cong 2 < w^2 > N .$$

Heisenberg[10] had proposed an effective diffusivity (K) in explaining the establishment of an inertial subrange (Hinze[11]) as

$$K = \int_{k_B}^{\infty} \left\{ \frac{E(k)}{k^3} \right\}^{1/2} dk .\qquad (8)$$

This reduces to

$$K_B \sim \frac{1}{4} \frac{\epsilon}{N^2} = \frac{1}{2} \frac{<w^2>}{N} . \qquad (9)$$

It is interesting to note here that there is a factor of approximately 1/2 introduced into the amplitude of the turbulent parameter ϵ_o via the definition of the turbulent spectrum Eq. (5). If instead we had used the rationale that the one-dimensional spectrum is only considered to the buoyant limit, then, Eq. (5) would read

$$E(k) = \alpha \epsilon^{2/3} k_B^{-2/3} ,$$

and Eqs. (7) and (9) would follow as

$$< w^2 > \sim \frac{\epsilon}{N} \text{ and } K_B \cong \frac{1}{2} \frac{<w^2>}{N} , \text{ respectively.}$$

3. DATA AND RESULTS

In this preliminary report, we examine the seasonal dependence of the vertical turbulent diffusivity (K) as a function of latitude and altitude, covering the latitude band $12°N$ to $71°N$. In a future report, we shall present in the same manner:

10. Heisenberg, W. (1948) Z. Physik, pp 124-628.
11. Hinze, J.O. (1959) Turbulence, McGraw-Hill, New York.

(1) occurrences for $R_i \leq 0.25$, (2) the resultant turbulent intensities $<w^2>$, and
(3) the rates of dissipation of turbulent kinetic energy (ϵ) over a larger latitudinal
region.

Figure 1a is a representative set of rawinsonde east-west wind component data.
The symbols are the discrete data values and the solid curve is the output of the Hermite
interpolation algorithm developed by Tsipouras and Cormier,[5] and here simply re-
ferred to as a "cubic spline fit". In Figure 1b the average shear taken from a linear fit to
the data values is compared to the derivative taken from the Hermite interpolation
algorithm, that is, using the derivative of the polynominal with the coefficients of
the fit. Using this technique, a close approximation to the average wind-component
gradients, or shears, and the temperature gradients are obtained at regular inter-
vals, by computer, without introducing problems associated with extremely large
or small numbers. The gradient Richardson number determined in this way, using
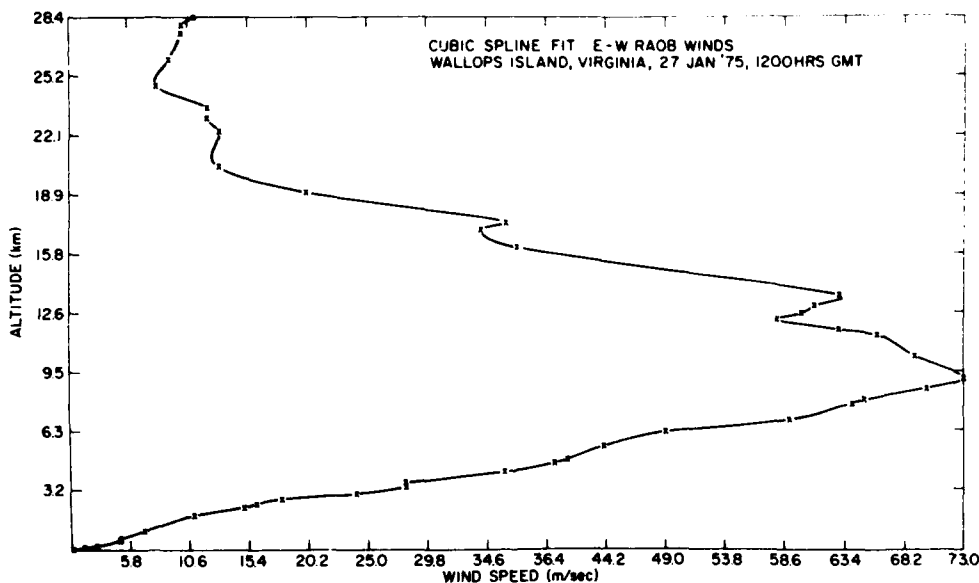the average gradients, provides only a relative measure of atmospheric stability
or instability.

Figure 1a. Cubic Spline Fit of the Rawinsonde East-West Wind Data. The solid
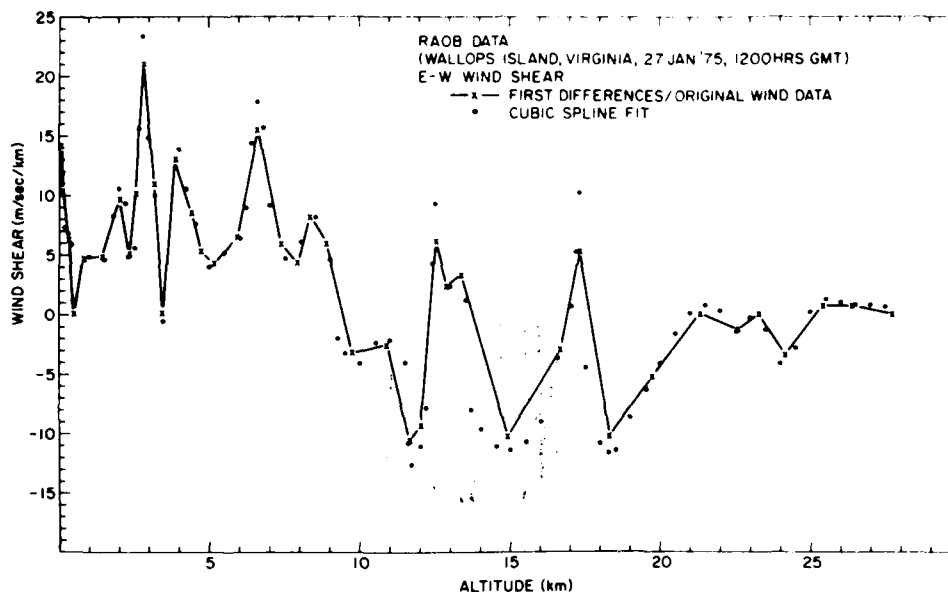line is the splined fit and the crosses (X) are the data points

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Figure 1b. The East-West Wind Shear. The linear calculations and the cubic spline fit are discussed in the text

The data base covers the years 1972 to 1976, and for this preliminary report we shall basically consider the analysis of the year 1976. The locations chosen are Barrow, Alaska (Lat 71°18' N), Yakutat, Alaska (59°31' N), Dayton, Ohio (39°52' N), Waycross, Georgia (31°15' N), Grand Cayman, B.W.1. (19°18' N), and San Andres, Columbia (12°35' N). Figures 2a, 2b, and 2c demonstrate the altitude distribution of turbulent diffusivities [Eq. (9)] for three of these sites for the spring and winter seasons from 1 km to ~ 30 km altitude. The data are calculated at 100-m intervals from the interpolated wind and temperature profiles and then averaged over 1-km intervals. What is immediately obvious is the striking reduction of turbulence at the tropopause (noted on the figures by the solid dots) and the significant enhancement of winter diffusivities over those of the spring season, particularly so in the stratosphere.

As observed for those latitudes when the tropopause is unambiguously determined (Figure 2a, San Andres 12°35' N), the rapid reduction of the turbulent diffusion coefficient at the tropopause altitude clearly delineates the disturbed region of the lower atmosphere from the more placid stratosphere. A similar phenomenon is observed at Barrow (70°18' N), but the tropopause is difficult to determine using the conventional logic. At Great Falls, there is again a marked reduction of turbulence at the tropopause, but significant turbulence in the stratosphere for the

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winter season. The spring season, on the other hand, at all sites shows very little turbulence in the stratosphere, even at the midlatutude sites.

The rapid reduction of turbulent diffusion is an unambiguous observation that suggests a more appropriate indicator of the tropopause would be the layer where there appears, in essence, a step function of a parameter involving both dynamical and thermal structure rather than a thermal demarcation alone. The turbulent diffusion coefficient, as calculated here, is such a parameter, clearly delineating a relatively placid region from the more disturbed region. The strong correspondence between the tropopause, when determinable, and the delineation altitude determined by the rapid cessation of turbulence is clearly displayed in Figures 3a and 3b. They are also contrasted to the isodensity contours of radioactive debris as reported by Danielson and Louis,[4] and it is clear that the tropopause, determined by the cessation of turbulence, shows a remarkable correspondence to that level of the atmosphere where the trace particles are rapidly depleted.
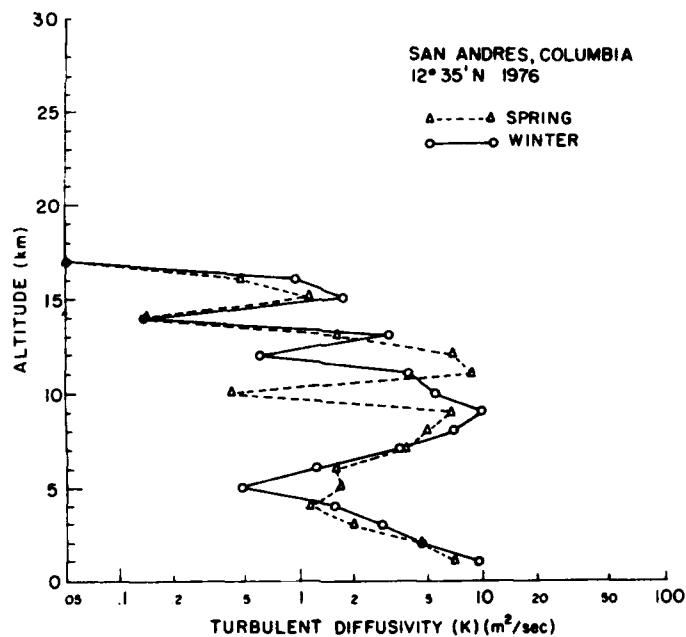

Figure 2a. One Kilometer Average of the Vertical Turbulent Diffusivities (K_{zz}) as Given by Equation (9) for the 1976 Winter and Spring Seasons at San Andres, Columbia
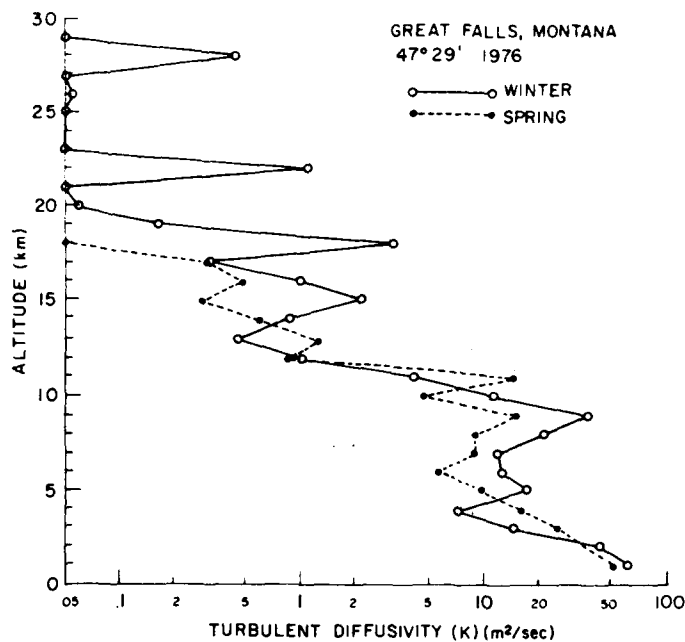
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Figure 2b. One Kilometer Average of the Vertical Turbulent Diffusivities (K_{zz}) as Given by Equation (9) for the 1976 Winter and Spring Seasons at Great Falls, Montana
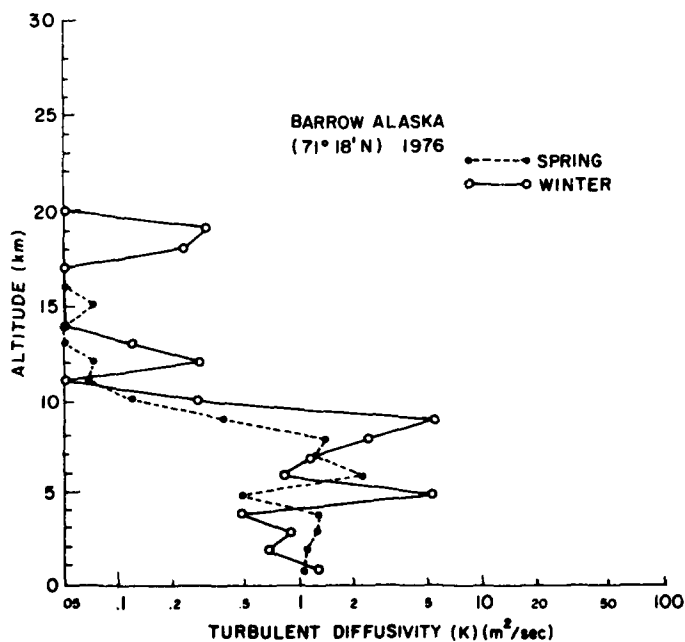

Figure 2c. One Kilometer Average of the Vertical Turbulent Diffusivities (K_{zz}) as Given by Equation (9) for the 1976 Winter and Spring Seasons at Barrow, Alaska
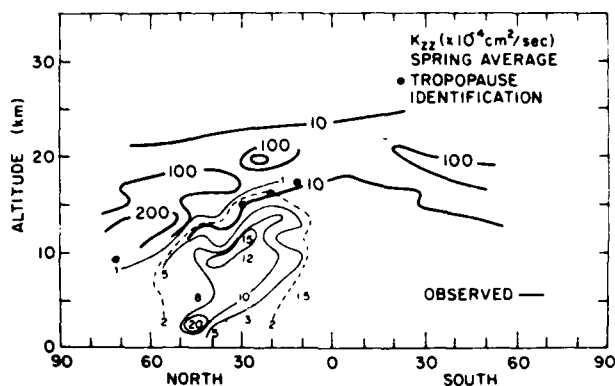
Figure 3a. Comparison of the Isodensity Contours
of Radioactive Debris (heavy solid lines) With the
Isocontours of Vertical Turbulent Diffusivities in
the Northern Hemisphere (light solid and dashed
lines). The numbers on the contours of turbulent
diffusivity are values expressed in $K_{zz} \times 10^{-4}$
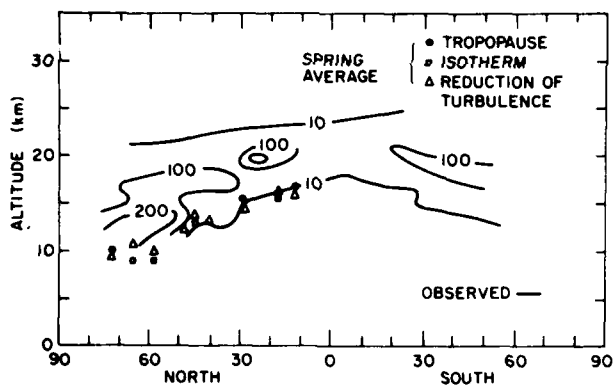cm^2/sec

Figure 3b. Comparison of the Isodensity Contours
With Three Parameters Marking the Tropopause

The large variability of stratospheric turbulence is further demonstrated in
Figure 4 where we observe a fairly intense increase of turbulence in the stratos-
phere for the winter season of 1976, as compared to that for 1975. The tropos-
pheric turbulence in 1975, on the other hand, is only slightly enhanced over that of
1976. This large variability of the stratospheric turbulent vertical diffusion above
the tropopause will have a significant effect upon seasonal and yearly variabilit, of
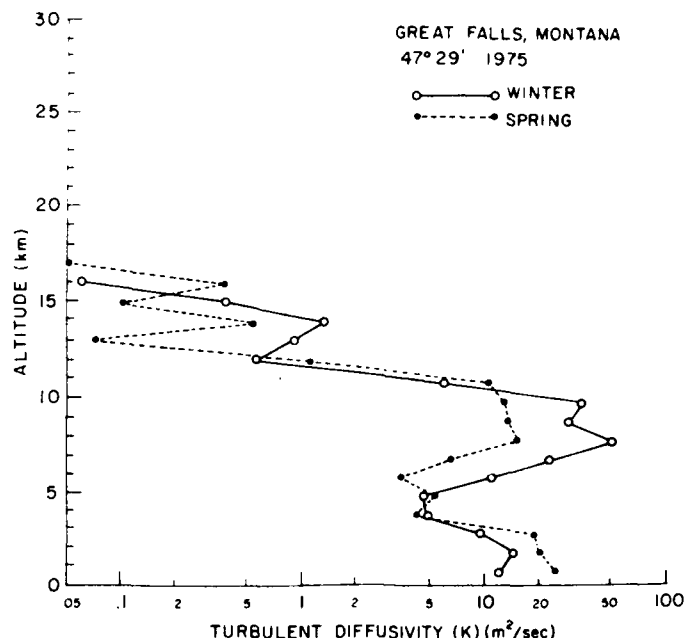transfer across the tropopause.

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Figure 4. The Vertical Turbulent Diffusivities of the 1975 Winter and Spring Seasons. These are for comparison with the 1976 results, Figure 2c, to demonstrate the large yearly differences observed in the stratosphere

Effective vertical transfer has been derived by the diffusion analysis of the stratospheric decay and tropospheric deposition of radioactive bomb debris.[1-3] Here the meridional transient eddies generated by Murakami[12] were utilized (References 1-3) to determine the meridional large-scale diffusivities, and then the continuity equation was used to estimate a vertical diffusivity that best approximated the observations cited by Danielson and Louis.[4] The comparison of their results with the measurements is given in Figure 5 (their Figure 9.11), and it is quite obvious that the derived slopes of the isodensity tracer contours are not at all commensurate with the measurements in the northern latitude. This conflict with the observations could be due to a number of facets, one being incorrect values of K_{yz}, the diagonal element of the meridional transfer function based upon the large-scale, heat-transfer calculation. Another is in their derived vertical diffusivity. This disagreement is evident in the comparison of our Figure 3a and

12. Murakami, T. (1962) Stratospheric Wind Temperature and Isobaric Height Conditions During the I.G.Y. Period, Part 1, M.I.T. Report No. 5

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their Table 9.1 (reproduced here as Figure 6) that shows the iso turbulent diffusion coefficients lines for $K \approx 1 \times 10^4$ cm^2/sec and $\approx 1.5 \times 10^4$ cm^2/sec. This later isocontour is also compared to the observations. There is no parallelism to the 10 PC$_i$/m^3 curve as this analysis displays, but a deeper penetration into the mid-latitude stratosphere with larger values than we determine. Our tropospheric values, on the other hand, are somewhat larger than theirs, approaching an order of magnitude difference when the similar averaging distance (3 km) is approximated (Figure 3a).
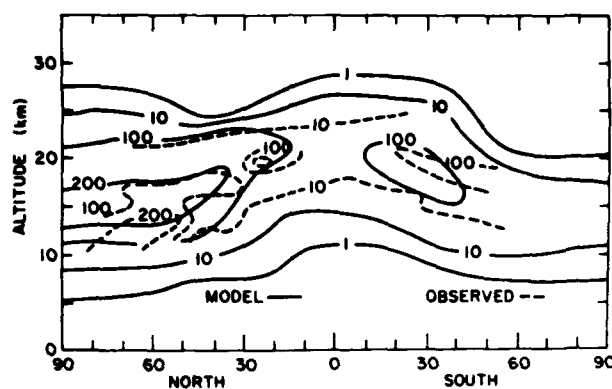

Figure 5. Observed Radioactivity Isodensity Contours (dashed lines) as Compared to the Model of Danielsen and Louis
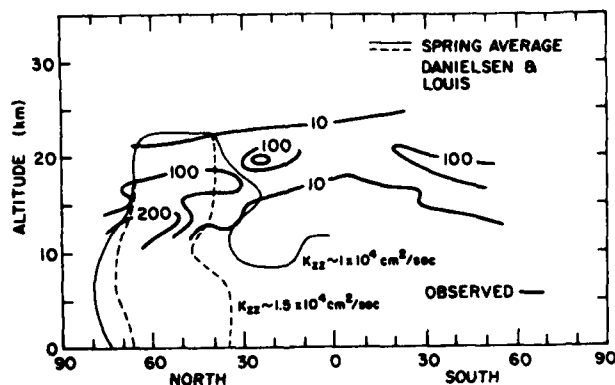

Figure 6. Spring Values of the Isocontours of the Vertical Turbulent Diffusivities (light solid and dashed lines) as Contrasted to the Observed Radioactivity Isodensity Contours

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4. CONCLUSION

In conclusion, we have demonstrated that vertical turbulent transfer, as determined from rawinsonde Richardson number calculations and utilization of boundary layer analysis, results in average turbulent diffusion coefficients in the tropopause and lower stratosphere. Comparison of iso-diffusion contours with radioactive debris isodensity contours demonstrates remarkable parallelism. Since turbulent diffusion operates upon species gradients (that is, $v_{diff} \approx K(\frac{1}{n}\frac{\partial n}{\partial z})$), we may infer that, indeed, the experimentally determined diffusivities dictate the maintenance of a flux perpendicular to the species isocontours and, thus, continued maintenance of these seasonal contours. Future efforts will be to extrapolate the results to southern latitudes and demonstrate other seasonal effects, and continue comparisons with other data.

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1. Reed, R. J., and German, K. E. (1965) A contribution to the problem of stratospheric diffusion by large-scale mixing, Mon. Weather Rev. 93:313-321.

2. Gudiksen, P. H., Fairhale, A. W., and Reed, R. J. (1968) Roles of mean meridional circulation and eddy diffusion in the transport of trace substance in the lower stratosphere, J. Geophys. Res. 73:4461-4473.

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12. Murakami, T. (1962) Stratospheric Wind Temperature and Isobaric Height Conditions During the I. G. Y. Period, Part 1, M. I. T. Report No. 5.

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